Chapter 3. WINTER HYDROLOGY

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3.1 Introduction

The winter hydrology component of the WEPP computer model is designed to simulate snow accumulation and density, snowmelt, and soil frost and thaw, all on a hourly basis. The snow accumulation routine predicts whether the hourly falling precipitation is rain or snow, as well as changes in snow depth and density. The melt component estimates the amount of snowmelt occurring for any given hour during the day. The frost component estimates the extent of frost development and thawing over the winter period as well as changes in soil water content and infiltration capacity of the soil during the winter period.

Winter hydrologic processes are an important part of the hydrology of watersheds located mainly north of 40 degrees latitude in the Northern hemisphere and south of 40 degrees latitude in the Southern hemisphere. Soil freezing and thawing influence the soil physical properties such as hydraulic conductivity, erodibility and soil water holding capacity. Freezing modifies the physical characteristics of soil, changing its ability to transmit or retain water (Benoit and Bornstein, 1970; Benoit and Mostaghimi, 1985; Campbell et al., 1970; Loch and Kay, 1978), its structural stability (Benoit, 1973; Mostaghimi et al., 1988), and its erodibility (Bisal and Nielsen, 1967). The development of soil frost is the result of complex interactions of several primary factors, including soil characteristics, type of tillage and residue management, surface roughness, type of vegetative cover, duration and extent of freezing temperatures, and the extent and timing of snow cover. The freezing process itself modifies those soil physical properties that, along with temperature, determine the depth and duration of soil frost. The magnitude of soil changes that takes place as a result of soil freezing depends on freezing temperature, soil water content at freezing, initial size of soil aggregates, and the number of freeze-thaw cycles that take place over winter. As a result, tillage-residue management combined with over winter frost action determines a soil's erodibility during winter thaw periods and from spring snowmelt to planting (Benoit et al., 1986).

Within the daily simulation mode, the winter routine is called if the following conditions prevail 1) a snowpack already exists 2) a soil frost layer already exists 3) average daily temperature is less than 0^oC. The winter hydrology routine works on a hourly basis, however, the WEPP climate input file provides daily values for precipitation, maximum and minimum temperature, dew point temperature and incoming radiation. Therefore, hourly precipitation, temperature and radiation need to be calculated before simulating snow accumulation or melt and frozen soil. In addition, the time of day when precipitation is occurring is needed in order to differentiate between snowfall and rainfall given hourly temperature.

3.2 Hourly Precipitation

For the case where the breakpoint precipitation option is selected in the climate input file, the amount of precipitation is calculated for each hour. The time when precipitation begins is available in this case. For the case when precipitation amount, duration of storm, time to peak/duration and ratio of maximum to average intensity option is selected, the disaggregation routine (Chapter 2) is used to determine the hourly precipitation. Furthermore, a random generation routine is used to estimate the time precipitation start.

3.3 Hourly Temperature

The daily minimum and maximum air temperature values are required input data. However, hourly air temperature and hourly temperature of the soil-snow-residue are needed for snowmelt and/or frost or thaw simulation. The equations to estimate the hourly distribution given daily temperature are taken from DeWit et al. (1978). In the WEPP model, we assume that the lowest temperature of the day occurs just before sunrise and the highest temperature of the day occurs at 2:00 pm. We also assume that between sunrise and 2:00 pm the temperature increases; otherwise the temperature decreases. Between sunrise and 2:00 pm, the hourly temperature is calculated using the following equation:

$$T_{hr} = T_{ave} - \frac{(T_{\max} - T_{\min})}{2} \cos\left[\frac{\pi(t_{hr} - 0.5 - t_{hrsr})}{14 - t_{hrsr}}\right] \qquad [3.3.1]$$

At all other times during the day, the hourly temperature is calculated using the following equation:

$$T_{hr} = T_{ave} + \frac{(T_{\max} - T_{\min})}{2} \cos\left[\frac{\pi\,(t_{hradj})}{10 + t_{hrsr}}\right] \qquad [3.3.2]$$

where T_{hr} is the hourly temperature (^oC), T_{ave} is the average air temperature of the day (^oC), $T_{\max}$ is the maximum air temperature of the day (^oC), $T_{\min}$ is the minimum air temperature of the day (^oC), t_{hrsr} is the sunrise hour (h), t_{hradj} is the adjusted hour and is equal to $t_{hr} - 0.5 - 14$ after 2:00 pm and it is equal to $t_{hr} - 0.5 + 10$ for the period before 2:00 pm, where t_{hr} is the current time of day (h). The average dew point temperature of the day is used to approximate the hourly dew point temperature, T_{hrdp} (^oC).

The model calculates the hourly adjusted surface temperature, T_{hra} (^oC), at the top surface of the residue-snow-frozen-layer system using a method suggested by Flerschinger et al. (1987). The hourly surface temperature is calculated using the following equation:

$$T_{hra} = \frac{R_{net} + (radco + conht\ (100\ v_{wind}))\ T_{ave}}{radco + conht\ (100\ v_{wind}) + \dfrac{efthco}{depth}} \qquad [3.3.3]$$

where T_{hra} is hourly surface temperature (^oC), R_{net} is net radiation ($Ly{\cdot}min^{-1}$), $conht$ is convective heat transfer coefficient ($Ly{\cdot}s{\cdot}min^{-1}{\cdot}cm^{-1}$), $radco$ is radiation coefficient ($Ly{\cdot}s{\cdot}min^{-1}{\cdot}cm^{-1}$), v_{wind} is wind velocity ($cm{\cdot}s^{-1}$), $efthco$ is effective system thermal conductivity ($L{\cdot}min^{-1}{\cdot}\,^oC^{-1}$), and $depth$ is the system depth (m).

Net radiation, R_{net}, is a calculated by the following equation:

$$R_{net} = R_{slp}\ (1.0 - alb) + (atem - suem)\ SBC\ T_{avek}^4 \qquad [3.3.4]$$

where R_{slp} is solar radiation on a sloping surface ($Ly{\cdot}min^{-1}$), alb is soil and/or snow albedo (%/100%), $atem$ is atmospheric emissivity, $suem$ is surface emissivity, SBC is the Stefan-Boltzman constant and is equal to $8.1247\,x\,10^{-11}\ Ly{\cdot}min^{-1}{\cdot}\,^oK^{-4}$ and T_{avek} is hourly air temperature (^oK). The surface emissivity, $suem$, for all layers is approximately 1.0. The radiation value is the same value of hourly radiation on a sloping surface that is calculated by the model (SUNMAP routine).

The surface albedo is dependent on the surface cover. If there is more than 0.5 centimeter of snow on the surface, then the albedo will be that of snow (0.5), otherwise it will be the same as the soil's albedo.

The atmospheric emissivity is a function of cloud cover and temperature.

$$atem = (1.0 - 0.84\ clouds)\ (1.0 - 0.261\ e^{x}) + 0.84\ clouds \qquad [3.3.5]$$

where *clouds* is the decimal fraction value for cloud cover, and x is equal to $0.000777 T_{ave}^2$.

The decimal fraction value for cloud cover is calculated using the following equation:

$$clouds = \frac{1.0 - \dfrac{R_{slp}}{R_{pot}}}{0.7} \qquad [3.3.6]$$

Both the radiation, R_{slp} ($MJ \cdot m^{-2}$), and potential radiation, R_{pot} ($MJ \cdot m^{-2}$), are daily values on the horizontal surface. This equation is based on the fact that clouds reflect approximately 70% of solar radiation and transmit only 30% to the earth's surface (Sutton, 1953). Both R_{slp} and R_{pot} are calculated in a separate subroutine based on the slope inclination, the slope facing direction or aspect, the calendar day, the predicted radiation from the climate input file, a solar constant and the latitude (Swift and Luxmoore, 1973). This method takes into account the effects of cloud cover and atmospheric transmissivity. The slope inclination, I, is calculated as:

$$I = arctan \left[\frac{S}{100} \right] \qquad [3.3.7]$$

where S is the land slope (%).

The convective heat transfer coefficient (*conht*) in equation 3.3.3 is calculated by the following equation:

$$conht = \frac{0.4^2\ \rho_{air}\ hca}{\log \left(\dfrac{U_z - disp + etr}{etr} \right)\ \log \left(\dfrac{U_z - disp + wsr}{wsr} \right)} \qquad [3.3.8]$$

where, 0.4 is the von Karman's constant, ρ_{air} is air density in $kg \cdot m^{-3}$ and is about 1.0, *hca* is heat capacity of the air and is equal to 101 $J \cdot kg^{-1} \cdot m^{-3}$, U_z is wind measurement height, usually 2 meters, *disp* is the zero plane displacement (m), *etr* is energy transfer roughness (m), and *wsr* is wind surface roughness (m). The value of zero plane displacement is a function of the canopy height and is calculated by the following equation:

$$disp = 0.77\ H_c \qquad [3.3.9]$$

when H_c is the canopy height (m).

The canopy cover data is available within the WEPP model. However, the wind surface roughness, *wsr*, could be any of several values. If there is no snow on the ground, the wind surface roughness is that of the canopy or soil surface if no canopy cover value is present. Otherwise if snow is present, then the roughness is equal to $(0.13 + H_c - D_{snew})$ where D_{snew} is the snow depth (m). If the snow is deeper than the canopy, then the wind surface roughness is set to 0.0002 meters. The energy transfer roughness, *etr* (m), is a function of wind surface roughness and is calculated using

$$etr = 0.2\ wsr \qquad [3.3.10]$$

When dealing with a layered system depth, the model calculates the gradient depth from the soil layer down to the first zero degree isotherm. The model also makes the assumption that if the frozen-layered system contains a top thaw layer that is greater than half of the top frost layer depth, it is all thawed and the gradient depth is set to 0. However, if a top thaw layer exists but it is less than half of the top frost layer depth, the model ignores that layer and treats it as all frozen. To calculate the heat transfer of a frozen-layered system, the model uses the harmonic mean for the layers in the system. The value of the harmonic mean is assigned to the variable surface thermal conductivity.

3.4 Aspect Adjustment

The aspect of a hillslope relative to the sun's angle which impinges upon it is calculated in the ASPECT subroutine. Values for slope steepness, slope aspect, latitude in radians, equivalent latitude, and the change in longitude with respect to equivalent slope and latitude are also calculated in this subroutine. The average slope of the Overland Flow Element (OFE) is calculated and converted to a decimal fraction and then to radians (*incrad*). The variable *radz* is the slope aspect in units of radians. These variables are then used to calculate the equivalent latitude and the change in longitude with respect to equivalent slope and latitude using the following equations:

$$eqla = arcsin\left[cos\,(incrad)\ sin\,(latrad)\ +\ sin\,(incrad)\ cos\,(latrad)\ cos\,(radz)\right] \qquad [3.4.1]$$

$$dellon = arctan\left[\frac{\sin(incrad)\ \sin(radz)}{\cos(incrad)\ \cos(latrad) - \sin(incrad)\ \sin(latrad)\ \cos(radz)}\right] \qquad [3.4.2]$$

where, *eqla* is equivalent latitude factor (*rad*), *incrad* is slope inclination (*rad*), *radz* is slope aspect (*rad*), *latrad* is latitude (*rad*), and *dellon* is change of longitude (*rad*). The values of *eqla* and *dellon* are used to adjust hourly radiation for the aspect effect.

3.5 Hourly Radiation

The method that calculates hourly radiation given the daily radiation is based on work by Swift and Luxmoore (1973) and Jensen et al. (1990). This calculation is done in the WEPP SUNMAP subroutine. This function returns the ratio between the potential radiation and the estimated radiation on a sloping surface for the day. This ratio distributes the daily value of solar radiation over a bell-shaped curve on an hourly basis.

The factor which represents the ratio of the radiation on the slope for the given hour to the total estimated radiation on the slope for the day is

$$RADF = \frac{720}{\pi}\ (SOLO)\ (RELD)\ cos\,(latrad)\ cos\,(SUND)\ \left[sin\,(HASR) - sin\,(HAST)\right]$$
$$+\ (HASR - HAST)\ sin\,(latrad)\ sin\,(SUND) \qquad [3.5.1]$$

where *SOLCO* is the solar constant (0.082 $MJ \cdot m^{-2} \cdot min^{-1}$), *RELD* is the relative distance of the earth from the sun (*rad*), *SUND* is the sun declination (*rad*), *HASR* is the position of the sun at sunrise (*rad*), and *HAST* is the position of the sun at sunset (*rad*). The relative distance of the earth from the sun (*RELD*) given the day of year is calculated by

$$RELD = 1.0 + 0.033 \cos \left(\frac{2\,\pi\, jday}{365} \right) \qquad [3.5.2]$$

where *jday* is the Julian day. The position of the sun at sunset (*HAST*) and sunrise (*HASR*) is calculated using

$$HAST = HAS - \frac{\pi}{24} \qquad [3.5.3]$$

$$HASR = HAS + \frac{\pi}{24} \qquad [3.5.4]$$

where *HAS* is the hour angle of the sun (*rad*). For a one hour time step, $HASR - HAST = \pi/12$. Based on the day of simulation the model calculates the sun's angle in the sky for each hour of the day. Hour angle of the sun (*HAS*) is calculated by

$$HAS = \left(t_{hr} + 0.06667\,(LS - L\,1) + STC - 12 \right) \frac{\pi}{12} \qquad [3.5.5]$$

where *LS* is longitude (*rad*) of the center of the local time zone, $L\,1$ is longitude of the site (*rad*). LS for the eastern U.S. time zone is 1.3090 radians, central U.S. time zone is 1.5708 radians, mountain U.S. time zone is 1.8326 radians, and pacific U.S. time zone is 2.0944 radians. *STC* is the solar time correction and is calculated in radians using the equation:

$$STC = 0.1645\, sin\,(2\,DF) - 0.1255\, cos\,(DF) - 0.025\, sin\,(DF) \qquad [3.5.6]$$

where *DF* is a day factor which represents the day of the simulation. Day factor is calculated by the following:

$$DF = \frac{2\,\pi\,(jday - 81)}{365} \qquad [3.5.7]$$

The radiation factor (*RADF*) is then used to calculate hourly radiation. The hourly estimated radiation (*hrrad*) on a sloping surface is estimated by multiplying the daily radiation on a sloping surface by the radiation factor. The variables calculated in the SUNMAP routine are potential horizontal radiation, R_{pot} ($MJ \cdot m^{-2}$), radiation on sloping surfaces on a hourly basis, *hrrad* ($MJ \cdot m^{-2}$), and half-day length (*h*). These values are used in the snowmelt calculations.

3.6 Snowmelt

The snowmelt calculations are performed in the MELT subroutine which is called from the main WINTER routine. The MELT routine is only called on hours of days when snow depth is greater than zero. The model calculates snowmelt and water melt separately to make sure that the snow density is 350 $kg \cdot m^{-3}$ before any snow melting is allowed to occur. If water melt is calculated, but the snow density is < 350 $kg \cdot m^{-3}$, no snowmelt is allowed to occur and the depth of the snowpack is decreased while the density of the snowpack is increased. In such a case, no water melt will be present because it all remains in the snowpack. If the density is less than but near 350 $kg \cdot m^{-3}$, and there is a significant amount of water melt, then the depth of the snowpack is decreased and the density is increased until it reaches 350 $kg \cdot m^{-3}$. Any water melt occurring after this point is passed to the DISAG routine for infiltration calculation. The *hrmelt* values are treated like breakpoint rainfall data. The water melt is recalculated based on the newly calculated snow depth and snow density. For this reason, the amount of melt calculated in the MELT subroutine and the final value of melt water for the hour (*hrmelt*) may not be the same.

The amount of water produced from a snowpack depends on the hourly weather and surface conditions. The equation to calculate the amount of water melt occurring for the hour-long period of time (*hrmelt*) is a modification of a generalized basin snowmelt equation for melt in open areas developed by the U.S. Army Corp of Engineers (1956, 1960). The equation was modified by Hendrick et al. (1971) to adapt it for use with readily available meteorological and environmental data. The Hendrick equation was further modified to make it compatible for use within the WEPP computer program.

$$hrmelt = 0.0254\ (amelt - bmelt + cmelt + dmelt) \qquad [3.6.1]$$

for $hrmelt \le D_{snew}$, where *hrmelt* is hourly melt water, 0.0254 is a conversion factor from inches to meters, and equations for calculating *amelt*, *bmelt*, *cmelt*, and *dmelt* follow. The first part, *amelt*, represents the hourly radiation (*hrrad*) component of the melt equation and is calculated using:

$$amelt = 0.0607\ hrrad\ (1.0 - cancov) \qquad [3.6.2]$$

where 0.0607 is a radiation conversion factor from *MJ* to *Ly*, *hrrad* is hourly calculated radiation ($MJ \cdot m^{-2}$), and *cancov* is canopy cover, (0-1). During hours of no sunlight, *amelt* will be zero. However, since some snowmelt can occur in direct solar radiation to about 3°C below freezing (Hendrick et al., 1971) therefore, when the hourly temperature (*hrtem*) is < 0.0 but > -4 .0, it is assumed that a small amount of melt will occur. This adjustment is made by the following equation.

$$amelt = amelt\ (0.36\ T_{hr} + 1.0) \qquad [3.6.3]$$

The second term in Eq. [3.6.1] is *bmelt*. This term represents the amount of long wave radiation coming down on the snowpack due to cloud cover and is calculated with:

$$bmelt = 0.84\ (1.0 - clouds) \qquad [3.6.4]$$

The third term, *cmelt*, is calculated by the following equation:

$$cmelt = 0.0188\ U\ (1.0–0.8\ cancov)\ (0.396\ T_{hr} + 1.404\ hrdew) \left[1.0– \left[\frac{1.0}{\log \left[\frac{U_z - disp + rough}{rough} \right]} \right] \right]$$

where, U is wind velocity ($m \cdot s^{-1}$), T_{hrdp} is hourly dew point temperature (^oC), U_z is the height of wind measurements (assume 2.0 meters), *disp* is the zero plane displacement of the surface (*m*), and rough is roughness of the surface (*m*). The roughness depends on snow depth, canopy or stubble height, or the soil's roughness, whichever is applicable.

$$disp = 0.77\ H_c \qquad [3.6.5]$$

where H_c is the canopy height (*m*), and represents the amount of canopy above the snow.

The final term, *dmelt*, represents the hourly air temperature and any heat input from rainfall and is calculated by the equation:

$$dmelt = T_{hr}\ (0.0382 + 0.014\ clouds) + T_{hr}\ (0.000496\ hrrain) \qquad [3.6.6]$$

where *hrrain* is hourly rainfall (m).

Equations 3.6.1-7 deal with four major energy components of the snowmelt process - temperature, radiation, vapor transfer, and precipitation. When calculating snowmelt, the following assumptions are made: 1) any precipitation that occurs on an hour when the maximum daily temperature is $< 0^oC$ is assumed to be snowfall; 2) no snowmelt will occur if the maximum daily temperature is $< -3^oC$; 3) the snowpack will not melt until the density of the snowpack is $\geq 350 kg \cdot m^{-3}$; 4) the surface soil temperature equals 0^oC during the melt period, 5) the temperature of a cloud base is approximately the same as the surface air temperature; and 6) the albedo of melting snow is approximately 0.5 (Sutton, 1953). Using Eq. [3.6.1], if the calculated value of snowmelt *hrmelt* is less than zero, then *hrmelt* is set to zero. If it is greater than the existing snow depth, D_{snew} (m), from the preceding hour, then *hrmelt* equals D_{snew}.

3.7 Snow Density

Events which change the snow density and therefore the depth of the snowpack are summarized here:

I. No snow on the ground
  - A. not snowing
  - B. is snowing

II. Snow on the ground
  - A. no melt has occurred
    - 1. no snow is falling
    - 2. ground drift
    - 3. daily settling
  - B. snow is falling
    - a. hourly snow
    - b. ground drift
    - c. falling drift
  - C. melt has occurred
    - 1. no snow is falling
      - a. snow density > 350, melt water
      - b. snow density ≤ 350, settled snow
    - 2. snow is falling

The variables that represent snow density are ρ_{sold}, (old value of snow density), and ρ_{snew} (newly calculated value of snow density); that is, before and after the snow depth has changed due to snowfall, melt and/or drift. The units of these variables are $kg \cdot m^{-3}$, and they are used interchangeably.

Since density of snow is a function of snow depth, the model also keep track of these depths both before and after any snowfall or drift has occurred. The variables D_{sold} refer to the snow depth before and D_{snew} represent the new snow depth after the snowfall or drifting has occurred. In the following section snow density and depth calculation for various conditions will be described.

The first condition is when there is no snow on the ground but it is snowing. Since the model assume all fresh snow has a density of 100 $kg \cdot m^{-3}$, and since there was no existing snow on the ground, we set ρ_{snew} equal to 100 $kg \cdot m^{-3}$. On the same time, we see that if no snow is on the ground and there is no falling snow, the density will simply be set to 0.0 $kg \cdot m^{-3}$.

The second condition is when the presence of snow has been recognized. The model checks for any snow that has melted during the past hour. If no melt had occurred, the model checks for snowfall. If no snow has fallen, then the only way that the depth could have changed is by either ground drifting or settling of the snowpack. Assuming this condition to be the case, the model then adds the amount of ground drifting to the variable D_{snew} to find the new depth and density. If ground drifting is the only factor which has changed the snow depth, we assume that the snow density to remain unchanged. Hence in ground drifting all of the drifting snow is from the existing snowpack. If no snow drifting has occurred, the settling of the snowpack is calculated using the equation:

$$setfa = 1 + e^{-2(fact)} \qquad [3.7.1]$$

The variable *setfa* is the hourly snow settling factor (0-1.0), and *fact* represents the number of days since snowfall has occurred. By multiplying the snow depth by *setfa* we can determine the new density of the settled snowpack for the hour. With this new settled density we can calculate the new depth of the settled snowpack by the following equation:

$$D_{snew} = \frac{D_{sold}\ \rho_{sold}}{\rho_{snew}} \qquad [3.7.2]$$

where, D_{snew} is the newly calculated depth of the settled snowpack (m), D_{sold} is the depth of the snowpack before settling(m), and ρ_{sold} is the old density before any settling ($kg \cdot m^{-3}$).

The other condition to consider is when no melting is occurring and it is snowing. In such a case, we add new snow to old snow if it exists, and calculate the new snow density using equation:

$$\rho_{snew} = \frac{\rho_{sold}\ (D_{sold} + grdri) + (100\ (hrsnow + faldr))}{D_{snew}} \qquad [3.7.3]$$

where ρ_{snew} is new snow density ($kg \cdot m^{-3}$), ρ_{sold} is the old snow density before addition of new snow ($kg \cdot m^{-3}$), D_{sold} is the depth before addition of new snow (m), *hrsnow* is hourly snowfall (m), and *faldr* is falling drift (m). The hourly snowfall and falling drift are multiplied by 100 $kg \cdot m^{-3}$, which is the density of fresh falling snow. D_{sold} and ground drift are multiplied by the original snow density, and D_{snew} represents the new snow depth after falling and drifting snow have been added.

The other condition is when snowmelt is occurring. The new snow depth after the hourly melt calculation is determine by subtracting snowmelt (*hrmelt*) from D_{sold}. If the new snow depth is 0.0, then all snow has melted and ρ_{snew} must go to 0.0. Otherwise, the model calculates the new snow density given the new snowpack depth by:

$$\rho_{snew} = \rho_{sold} \frac{D_{sold}}{D_{snew}} \qquad [3.7.4]$$

Since the model calculates the new density after the melt calculation, we must make sure that the new snow density is greater than 350 $kg{\cdot}m^{-3}$ for melting to occur. Up until that value, the melting snow does not reach the soil surface but simply increases the snowpack's density. If ρ_{snew} < 350 $kg{\cdot}m^{-3}$, the amount of melted snow and water is 0.0.

The last condition is the case when melting is occurring at the same time as snowfall. In such a situation, the model adds the amount of hourly snow to the snow depth and calculates the new density using the following:

$$\rho_{snew} = \frac{(\rho_{sold}\ D_{sold)} + (100\ hrsnow)}{D_{snew}} \qquad [3.7.5]$$

The WEPP model sets an upper limit on snow density of 522 $kg{\cdot}m^{-3}$.

3.8 Frost Simulation

The soil frost subroutine is based on simple heat flow theory. It assumes that heat flow in a frozen or unfrozen soil or soil-snow system is unidirectional and that the average 24 hour temperature of the system surface-air interface is approximated by average daily air temperature. The FROST subroutine predicts hourly frost and thaw development for various combinations of snow, residue, and tilled and/or untilled soil and is driven only by daily inputs of maximum and minimum air temperature and snow depth. Snow and soil thermal conductivity and water flow components are considered as constants. The subroutine yields values for hourly frost depth, thaw depth, number of freeze-thaw cycles, water accumulated in frozen soil, and infiltration capacity of the tilled layer or top 0.20 meters of soil if the soil is untilled. The soil frost subroutine operates by a hourly bookkeeping process that compares calories of heat lost or gained at the soil surface to heat flow from deeper unfrozen soil layers. Net calories of heat lost or gained are converted to meters of frozen or thawed soil. Uni-directional heat flow through the frozen soil or soil-residue-snow system is calculated from the relation:

$$Q_{srf} = \frac{K_{srf}\ \Delta T_{srf}}{Z_{srf}} \qquad [3.8.1]$$

where Q_{srf} is the heat flux through the snow-residue-frozen soil system ($W{\cdot}m^{-2}$), K_{srf} is the average thermal conductivity through the combined snow residue-frozen soil depth thickness ($W{\cdot}m^{-1}{\cdot}{}^oC^{-1}$), ΔT_{srf} is the temperature difference across the snow-residue-frozen soil thickness (^oC), and Z_{srf} is the depth or thickness of the combined snow-residue-frozen soil layer (m).

Thus, heat flow through the snow-residue-frozen soil layer is the product of an average thermal conductivity for the layer and an average temperature gradient, with the gradient being the difference between average daily air temperature and the zero degree isotherm at the bottom of the frozen soil.

The basic assumption is made that the average temperature of the soil (snow) - air interface over a 24 hour period is equal to the average air temperature for the same period. The validity of this assumption varies with location as a function of those items such as emissivity, radiation, cloud cover, and wind. For this reason, the hourly surface temperature that drives the frost subroutine is computed by a surface energy balance routine that modifies hourly air temperature by a local accounting of wind speed, solar radiation, cloud cover, and atmospheric emissivity (Flerschinger, 1987).

The average thermal conductivity for a layered system can be shown to equal the harmonic mean for the layers in the system and is given by:

$$K_{srf} = \frac{Z_{srf}}{\sum_{i=1}^{N} \left(\frac{Z_i}{K_i} \right)} \qquad [3.8.2]$$

where Z_i is the thickness of each layer (m), K_i is the thermal conductivity of each layer, ($W\ m^{-1}\,{}^{\circ}C^{-1}$) N is the number of layers, and Z_{srf} is the sum of the individual layer thicknesses.

The soil frost subroutine is designed to handle a system with up to four layers - snow, residue, tilled soil, and untilled soil. In this case the average thermal conductivity equation becomes:

$$K_{srf} = \frac{(K_{snow}\,K_{res}\,K_{ftill}\,K_{futil})(S_{nowd} + R_{esd} + T_{illd} + U_{tilld})}{K_{snow}\,K_{res}\,K_{ftill}\,U_{tilld} + K_{snow}\,K_{res}\,K_{futil}\,T_{illd} + K_{snow}\,K_{ftill}\,K_{futil}\,R_{esd} + K_{res}\,K_{ftill}\,K_{futil}\,S_{nowd}} \qquad [3.8.3]$$

where K_{snow} is the thermal conductivity of snow ($W{\cdot}m^{-1}{\cdot}{}^{o}C^{-1}$) K_{res} is the thermal conductivity of residue ($W{\cdot}m^{-1}{\cdot}{}^{o}C^{-1}$), K_{ftill} is the thermal conductivity of frozen tilled soil ($W{\cdot}m^{-1}{\cdot}{}^{o}C^{-1}$), K_{futil} is the thermal conductivity of frozen untilled soil ($W{\cdot}m^{-1}{\cdot}{}^{o}C^{-1}$), S_{nowd} is the snow depth (m), R_{esd} is the residue thickness (m), T_{illd} is the tilled soil depth (m), and U_{tilld} is the untilled soil depth (m). With this approach, if any or all of the snow, residue or tilled depths are zero, the thermal conductivity reduces to the harmonic mean of the remaining layers.

Over any 24 hour period, Q_{srf} must be balanced by heat flow Q_{uf} from the unfrozen soil below the frozen layer. The frost subroutine defines Q_{uf} as the sum of heat transferred by the thermal conductivity properties of the soil matrix, the latent heat of fusion in freezing transferred water, and losses in heat content of the soil. That is:

$$Q_{uf} = K_{uf} \left(\frac{T_{uf}}{Z_{uf}} \right) + LK_w \left(\frac{P}{Z_{uf}} \right) + C_{uf}\, dT_{uf}\, Z_c \qquad [3.8.4]$$

where Q_{uf} is the heat flow from unfrozen soil ($W{\cdot}m^{-2}$), K_{uf} is the thermal conductivity of unfrozen soil ($W{\cdot}m^{-1}{\cdot}{}^{o}C^{-1}$), T_{uf} is the change in temperature from the 0 degree isotherm to the depth of stable temperature (${}^{o}C$), Z_{uf} is the depth of unfrozen soil to the point of stable temperature (m), L is the latent heat of fusion ($W{\cdot}s{\cdot}m^{-3}$), K_w is the unsaturated hydraulic conductivity of soil ($m{\cdot}s^{-1}$), P is the change in total water potential (m), C_{uf} is the heat capacity of the unfrozen soil ($W{\cdot}m^{-3}{\cdot}{}^{o}C^{-1}$), dT_{uf} is the change in temperature of a unit volume of soil in unit time (${}^{o}C$), and Z_c is the depth of unfrozen soil that supplies heat as a result of changes in soil temperature (m) (WEPP assumes a constant value of 1.0 meter).

In this equation, the soil temperature and water potential gradients are those that exist just below the 0 degree isotherm. As a practical convenience, the model assumes that heat flow through soil thermal conductivity and soil water movement are separate and discrete units of heat transfer.

The subroutine operates by iteratively balancing over each 24 hour period the heat lost through the snow-residue-frozen soil zone with heat flow through the unfrozen soil to the freezing front. Iteration is on an hourly basis for each 24 hour period. During the balancing process, it is assumed that heat lost through the frozen zone is first balanced by heat flow in the unfrozen soil as a result of the soil's temperature gradient and thermal conductivity. Additional heat loss is balanced by the heat of fusion released by freezing water that is held in place or migrates to the freezing front. Further heat loss is balanced by changes in soil heat content of the unfrozen soil, the magnitude of which is computed by difference.

3.9 Snow Drifting

The snowdrift subroutine determines the distribution of snow over the hillslope by estimating the depth of snow on the ground at the end of a day in any slope section, depending on the weather that day and the topography. The snow drifting equations described here are not currently active in the version of the WEPP computer program released in August, 1995, but they are expected to be used in future versions. Calculations are based on several initial assumptions:

— the threshold wind velocity for moving falling snow ≈0.89 $m\ s^{-1}$ measured at a height of 2 m,
— the surface roughness of a uniform snowpack 0.0002 m, and
— the snow storage capacity of a tilled layer = the random roughness.

The amount of snow trapped and stored by standing vegetation is the storage capacity, S_t, and is a function of the height and the projected stem area, or basal density, of the vegetation, the surface roughness, and the amount of standing biomass. S_t is calculated as:

$$S_t = \varepsilon\, H\, d_b\, \frac{R_t}{R_o} + z_o \qquad [3.9.1]$$

where S_t is the storage capacity of snow (m), ε is the a trapping efficiency (%), H is the height of standing vegetation (m), d_b is the basal density of standing vegetation ($m \cdot m^{-1}$), R_t is the standing residue mass after tillage ($kg \cdot ha^{-1}$), R_o is the standing residue mass before tillage ($kg \cdot ha^{-1}$), and z_o is the surface random roughness (m). The trapping efficiency, ε, reflects the effect of vegetative height and is calculated by:

$$\varepsilon = (e^{-0.1H}) - 0.1 \qquad [3.9.2]$$

The basal density of the standing vegetation, d_b, is a function of the mean diameter and the plant population and is calculated by:

$$d_b = \frac{d_s\, p_o^{1/2}}{25} \qquad [3.9.3]$$

where d_s is the mean stem diameter of standing vegetation (m), and p_o is the plant population ($plants \cdot ha^{-1}$).

User inputs to the subroutine consist of the slope facing direction (A) in degrees from north, the land slope (S) in percent, and the length (L) and width (W) of the slope section in meters. The surface random roughness is obtained from the SOIL subroutine, and precipitation (P_d), mean minimum daily temperature ($T_{\min}$), mean daily wind speed (v_2), and mean daily wind direction (W) are all obtained from the climate input file. The height of standing residue (H), mean stem diameter (d_s), plant population (P_o), and the standing residue mass before and after tillage (R_o and R_t) are all obtained from the plant growth and residue decomposition components.

The snowdrift subroutine works in two parts, first calculating the amount of scouring or drifting of newly-falling snow occurring on a slope section, or plane, during the day, including any snow drifting into the section from an upwind section, and then calculating the amount of drifting or scouring of the existing snowpack. If the drift rate (D_r) calculated for an upwind section is negative, indicating that snow in that section is drifting out of the section (scouring), then the amount of snow scoured from that upwind section is added to the snow available for movement in the next downslope section. For falling snow, a threshold velocity of 0.89 $m\ s^{-1}$ at a height of 2 m is assumed for the incipient blowing of snow. In order to route the blowing snow across the slope, certain assumptions must be made. An upwind slope section at the top of the hillslope must accumulate snow unless the wind is blowing in a direction directly perpendicular to the direction in which the slope faces. Snow drifting onto the upslope section from upwind is distributed onto successive downwind sections according to a decay function. A downwind slope section at the top of the hillslope must scour unless the wind is blowing directly perpendicular to the direction in which the slope faces. If the wind blows perpendicular to a slope section, then no scouring or drifting occurs and the net change in snow accumulation in that section due to wind is zero. It is also assumed that there is always a sufficient supply of snow available to satisfy a drifting requirement for a given slope section.

The friction velocity at the snow surface is calculated using a commonly used mathematical representation of the wind profile (Schlichting, 1979):

$$v_h = \left(\frac{v_*}{k} \right) \ln \left(\frac{h}{z_o} \right) \qquad [3.9.4]$$

where v_h is the wind velocity measured at height h ($m \cdot s^{-1}$), v_* is the friction velocity at the snow surface ($m \cdot s^{-1}$), k is the von Karman's constant (assumed to be 0.4), h is the height above the surface (m), and z_o is the surface roughness (m).

After v_* is determined, if the value of v_* is < 0.087 $m \cdot s^{-1}$ (the friction velocity corresponding to a wind velocity of 0.89 $m \cdot s^{-1}$ measured at a height of 2 meters), then no movement of falling snow will occur and the new snow depth will be equal to the snow depth from the preceding day plus the depth of new snowfall. If $v_* \geq 0.087$ $m \cdot s^{-1}$, falling snow will begin to drift and the transport capacity of the wind is calculated from an equation developed by Bagnold (1941) and modified by Iversen et al. (1975):

$$q_{sf} = c \left(\frac{\rho_{air}}{g} \right) \left(\frac{v_f}{v_{thf}} \right) (v_*^2)(v_* - v_{thf}) \qquad [3.9.5]$$

where q_s is the transport rate of snow ($kg \cdot m^{-1} \cdot s^{-1}$), c is a proportionality constant (= 100), ρ_{air} is the density of air ($kg\ m^{-3}$), g is the acceleration of gravity ($m \cdot s^{-2}$), v_f is the settling velocity of a snow particle ($m \cdot s^{-1}$) (for falling snow assume 0.35 $m \cdot s^{-1}$ for a 0.150 mm snow particle falling in still air) (Schmidt, 1982), v_{thf} is the threshold velocity for incipient motion of falling snow ($m \cdot s^{-1}$) (assume 0.087 $m \cdot s^{-1}$), and v_* is the friction velocity at the snow surface ($m \cdot s^{-1}$).

The drift rate of falling snow is then determined from:

$$D_f = 86.4 \frac{q_{sf}}{d_f L_p} \qquad [3.9.6]$$

where D_f is the drift rate of falling snow ($m \cdot d^{-1}$), d_f is the density of falling snow ($kg \cdot m^{-3}$) (≈100 $kg \cdot m^{-3}$), and L_p is the distance across a slope section parallel to the wind direction (m).

While the threshold velocity for incipient movement of snow varies with the nature of the snow surface (Radok, 1977), for a uniform surface of freshly fallen snow, the threshold friction velocity for movement of snow from the snowpack is approximately 0.25 $m\ s^-1$ (Tabler and Schmidt, 1986) which is equivalent to a wind velocity of about 5.76 $m \cdot s^{-1}$ at a height of 2 meters. However, the threshold friction velocity for movement of snow from a snowpack is a function of the density of the snowpack and, thus, will increase with time since deposition (Schmidt, 1980). The threshold velocity for movement of snow from a snowpack, v_{thg} ($m \cdot s^{-1}$), can be estimated from

$$v_{thg} = \frac{-0.023\sqrt{\dfrac{z_o}{3.2x10^{-7}}}}{\left\{1 - \sin\left[\tan^{-1}\left(\dfrac{S}{100}\right)\right]\right\}\ln(0.001\, dg)} \quad [3.9.7]$$

where z_o is the surface roughness (m), S is the slope (%), and ρ_{sp} is the density of the snowpack ($kg{\cdot}m^{-3}$). If the calculated value of $v_* \geq v_{thg}$, then snow on the ground will begin to move. The transport capacity of the wind for moving ground snow is then calculated in a fashion similar to that for calculating the transport capacity of the wind for moving falling snow, as:

$$q_{sg} = c\left(\frac{\rho_{air}}{g}\right)\left(\frac{v_g}{v_{thg}}\right)(v_*^2)(v_* - v_{thg}) \quad [3.9.8]$$

where q_{sg} is the transport rate of ground snow ($kg{\cdot}m^{-1}{\cdot}s^{-1}$), and v_g is the settling velocity of a ground snow particle ($m{\cdot}s^{-1}$) (for ground snow assume 0.75 $m{\cdot}s^{-1}$ for a 0.220 mm ice sphere falling in still air) (Schmidt, 1982), and v_{thg} is the threshold velocity for incipient motion of ground snow ($m{\cdot}s^{-1}$). The drift rate of ground snow is then calculated from:

$$D_g = \frac{8.64\, q_{sg}}{\rho_{sp} L_p} \quad [3.9.9]$$

where D_g is the drift rate of ground snow ($m{\cdot}d^{-1}$), and ρ_{sp} is the density of the snowpack ($kg{\cdot}m^{-3}$).

The density of a snowpack on the ground is a function of several factors, including time and temperature. Daily changes in the density of the snowpack are calculated on the basis of the initial depth of the snowpack and how much snowmelt occurs each day. In the absence of snowmelt, changes in snowpack density are estimated daily from the relationship:

$$\rho_{sp} = 0.522 - \left(\frac{20.5}{D}\right)(1 - e^{-0.0148D}) \quad [3.9.10]$$

where D is the existing snow depth (m). This relationship is based on 14 years of pre-melt snowdrift data (Tabler, 1985). If snowmelt occurs while the snowpack density is less than 350 $kg{\cdot}m^{-3}$, the depth of the snowpack is reduced by the amount of the melt but the amount of melt water is added to the remaining snowpack, thus, increasing its density. Once the density of the snowpack equals or exceeds 350 $kg\ m^{-3}$, any additional melt water will either infiltrate the ground or run off.

If D_g exceeds the existing snow depth, D, then D_g is set equal to D. The total drift rate, D_r, is the sum of the drift rates for falling snow and ground snow, or:

$$D_r = D_f + D_{g.} \quad [3.9.11]$$

After the total drift rate is calculated based on the transport capacity of the wind to carry snow, the direction of the wind with respect to the direction the slope faces will determine whether the snow is drifting into the slope section or out of it. Maximum scouring will occur if the direction from which the wind is blowing is the same as the direction the slope is facing and maximum drifting will occur if the slope facing direction and the wind direction are exactly opposite each other. As previously stated, if the wind is blowing perpendicular to the slope, net scouring or drifting will be zero. Thus, to determine the net movement of blowing snow into or out of an area, the total drift rate D_r must be multiplied by a factor, s_c, to reflect either scouring or drifting:

$$s_c = 0.0111\ |A - W| - 1.0 \qquad [3.9.12]$$

where A is the slope azimuth (degrees from north), and W is the wind direction (degrees from north).

As the degree of slope inclination increases, the efficiency of the drifting process tends to decrease. This is accounted for by multiplying the net movement of blowing snow by an efficiency factor, i, based on land slope:

$$i = 1 - \sin\left[arctan \left(\frac{S}{100} \right) \right] \qquad [3.9.13]$$

If, due to wind angle and slope azimuth, the net movement of snow is positive, i.e. drifting into a slope section rather than out of it, the drifting snow will be distributed along the slope in a downwind direction. The amount of drifting snow falling on any slope section can be approximated by an exponential decay function:

$$D_p = 1 - \left(\frac{e^{-L_r}}{1 + 10\ L_r} \right) \qquad [3.9.14]$$

where D_p is the total percentage of available drifting snow falling on an upslope area (%) and L_r is the ratio of the length of the upslope area to the total slope length.

Not all of the moving snow will be deposited since some of it will evaporate. Net sublimation or evaporation losses can be an important consideration in climatic hydrological balances (Branton et al., 1972). The amount of evaporation is a function of air temperature, relative humidity, solar radiation, and particle diameter (Sturges and Tabler, 1981). An estimate of the amount of evaporation occurring can be obtain ed by considering the distance the snow is being blown along a slope and assuming that under average conditions, complete evaporation would occur after being blown a distance of about 3050 m (Tabler, 1975). Then:

$$D_r = D_r\ e^{-0.00066\ L_p} \qquad [3.9.15]$$

where L_p is the distance across a slope section parallel to the wind direction (m). Evaporation losses are only calculated for those sections in which drifting is occurring. Evaporation losses of snow from areas that are scouring would be accounted for in their downwind areas and are neglected here.

3.10 References

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3.11 List of Symbols

Symbol	Description	Units
A	slope facing direction in degrees from North	$\circ$
alb	soil and/or snow albedo	%/100%
C_{uf}	heat capacity of the unfrozen soil	$W \cdot m^{-3} \cdot {}^oC^{-1}$
c	proportionality constant	-
$cancov$	canopy cover	0-1.00-1.0
$clouds$	the decimal fraction value for cloud cover	*Fraction*
$conht$	convective heat transfer coefficient	$Ly \cdot s \cdot min^{-1} \cdot cm^{-1}$
ΔT_{srf}	temperature difference across the snow-residue-frozen soil thickness	${}^oC$
D_f	drift rate of falling snow	$m \cdot d^{-1}$
D_g	drift rate of ground snow	$m \cdot d^{-1}$
D_p	total percentage of available drifting snow falling on an upslope area	%
D_r	total drift rate	$m \cdot d^{-1}$
D_{snew}	newly calculated depth of settled snowpack	m
D_{sold}	depth of the snowpack before settling	m
DF	day factor which represents the day of the simulation	-
d_b	basal density of standing vegetation	$m \cdot m^{-1}$
d_f	density of falling snow	$kg \cdot m^{-3}$
d_s	mean stem diameter	m
dT_{uf}	change in temperature of a unit volume of soil in unit time	${}^oC$
$dellon$	change of longitude	rad
$depth$	the system depth	m
$disp$	zero plane displacement	m
$dmelt$	hourly air temperature and any heat input from rainfall	-
ε	a trapping efficiency	%
$efthco$	effective system thermal conductivity	$L \cdot min^{-1} \cdot {}^oC^{-1}$
etr	energy transfer roughness	m
$eqla$	equivalent latitude factor	rad
$fact$	number of days since snowfall has occurred	d
$faldr$	falling drift	m
g	acceleration of gravity	$m \cdot s^{-2}$
H	height of standing vegetation	m
H_c	canopy height	m
HAS	hour angle of the sun	rad
$HASR$	position of the sun at sunrise	rad
$HAST$	position of the sun at sunset	rad
h	height above the surface	m
hca	heat capacity of the air	$J \cdot kg^{-1} \cdot m^{-3}$
$hrmelt$	hourly melt water	m

$hrrad$	hourly estimated radiation	$MJ{\cdot}m^{-2}$
$hrrain$	hourly rainfall	m
$hrsnow$	hourly snowfall	m
$hrtem$	hourly temperature	^{o}C
$incrad$	slope inclination	rad
$jday$	Julian day	$julian\ day$
K_{ftill}	thermal conductivity of frozen tilled soil	$W{\cdot}m^{-1}{\cdot}^{o}C^{-1}$
K_{futil}	thermal conductivity of frozen untilled soil	$W{\cdot}m^{-1}{\cdot}^{o}C^{-1}$
K_i	thermal conductivity of each layer	$W{\cdot}m^{-1}{\cdot}^{o}C^{-1}$
K_{res}	thermal conductivity of residue	$W{\cdot}m^{-1}{\cdot}^{o}C^{-1}$
K_{snow}	thermal conductivity of snow	$W{\cdot}m^{-1}{\cdot}^{o}C^{-1}$
K_{srf}	average thermal conductivity through combined snow-residue-frozen soil depth thickness	$W{\cdot}m^{-1}{\cdot}^{o}C^{-1}$
K_{uf}	thermal conductivity of unfrozen soil	$W{\cdot}m^{-1}{\cdot}^{o}C^{-1}$
K_w	unsaturated hydraulic conductivity of soil	$m{\cdot}s^{-1}$
k	von Karman’s constant	-
L	latent heat of fusion	$W{\cdot}s{\cdot}m^{-3}$
$L1$	longitude of the site	rad
LS	longitude	rad
L_r	ratio of the length of the upslope area to the total slope length	$m{\cdot}m^{-1}$
L_p	distance across a slope section parallel to the wind direction	m
$latrad$	latitude	rad
P	change in total water potential	m
P_d	daily precipitation	m
Q_{srf}	heat flux through the snow-residue-frozen soil system	$W{\cdot}m^{-2}$
Q_{uf}	thermal conductivity of unfrozen soil	$W{\cdot}m^{-2}$
q_s	transport rate of snow	$kg{\cdot}m^{-1}{\cdot}s^{-1}$
q_{sg}	the transport rate of ground snow	$kg{\cdot}m^{-1}{\cdot}s^{-1}$
R_{esd}	residue thickness	m
R_o	standing residue mass before tillage	$kg{\cdot}ha^{-1}$
R_{slp}	solar radiation on a sloping surface	$Ly{\cdot}min^{-1}$
R_t	standing residue mass after tillage	$kg{\cdot}ha^{-1}$
R_{net}	net radiation	$Ly{\cdot}min^{-1}$
R_{pot}	potential radiation on a horizontal surface	$MJ{\cdot}m^{-2}$
$RADF$	radiation factor	-
$RELD$	relative distance of the earth from the sun	rad
$radco$	radiation coefficient	$Ly{\cdot}s{\cdot}min^{-1}{\cdot}cm^{-1}$
$radz$	slope aspect	rad
ρ_{air}	air density	$kg{\cdot}m^{-3}$
ρ_{sold}	old density before any settling	$kg{\cdot}m^{-3}$
ρ_{snew}	new snow density	$kg{\cdot}m^{-3}$
ρ_{sp}	density of the snowpack	$kg{\cdot}m^{-3}$
$rough$	roughness of surface	m
S	land slope	%
S_{nowd}	snow depth	m
S_t	storage capacity of snow	m
SBC	Stefan-Boltzman constant	$Ly{\cdot}min^{-1}{\cdot}^{o}K^{-4}$
$SOLCO$	solar constant	$MJ{\cdot}m^{-2}{\cdot}min^{-1}$
STC	solar time correction	rad
$SUND$	sun declination	rad
$setfa$	hourly snow settling factor	0-1.0

T_{ave}	average air temperature of the day	^oC
T_{avek}	hourly air temperature	^oK
T_{hr}	hourly temperature	^oC
$T_{\max}$	maximum air temperature of the day	^oC
$T_{\min}$	minimum air temperature of the day	^oC
T_{hra}	hourly surface temperature	^oC
T_{hrdp}	hourly dew point temperature	^oC
T_{illd}	tilled soil depth	m
T_{uf}	change in temperature from the 0 degree isotherm to the depth of stable temperature	^oC
t_{hr}	current time of day	h
t_{hradj}	adjusted hour	h
t_{hrsr}	sunrise hour	h
U_{tilld}	untilled soil depth	m
U	wind velocity	$m{\cdot}s^{-1}$
U_z	height of wind measurements	m
v_2	mean daily wind speed	$m{\cdot}s^{-1}$
v_*	friction velocity at snow surface	$m{\cdot}s^{-1}$
v_f	settling velocity of a snow particle	$m{\cdot}s^{-1}$
v_g	settling velocity of a ground snow particle	$m{\cdot}s^{-1}$
v_h	wind velocity measured at height h	$m{\cdot}s^{-1}$
v_{thg}	threshold velocity for incipient motion of ground snow	$m{\cdot}s^{-1}$
v_{thf}	threshold velocity for incipient motion of falling snow	$m{\cdot}s^{-1}$
v_{wind}	wind velocity	$cm{\cdot}s^{-1}$
W	width of slope section	m
W	mean daily wind direction	rad
wsr	wind surface roughness	m
x	equal to $0.000777T_{ave}^2$	-
Z_c	depth of unfrozen soil that supplies heat as a result of changes in soil temperature	m m
Z_i	thickness of each layer	m
Z_{srf}	depth or thickness of combined snow-residue-frozen soil layer	m
Z_{uf}	depth of unfrozen soil to the point of stable temperature	m
z_o	surface random roughness	m